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Global patterns and dynamics of climate-groundwater interactions

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28 **Summary Paragraph**

29 Groundwater is the largest available store of global freshwater¹, upon which more than two
30 billion people rely². It is therefore important to quantify the spatiotemporal interactions
31 between groundwater and climate. However, current understanding of the global scale
32 sensitivity of groundwater systems to climate change^{3,4} – as well as the resulting variation in
33 feedbacks from groundwater to the climate system^{5,6} - is limited. Here, using groundwater
34 model results in combination with hydrologic datasets, we examine the dynamic timescales
35 of groundwater system responses to climate change. We show that nearly half of global
36 groundwater fluxes could equilibrate with recharge variations due to climate change on
37 human (~100 year) timescales, and that areas where water tables are most sensitive to
38 changes in recharge are also those that have the longest groundwater response times. In
39 particular, groundwater fluxes in arid regions are shown to be less responsive to climate
40 variability than in humid regions. Adaptation strategies must therefore account for the
41 hydraulic memory of groundwater systems which can buffer climate change impacts on water
42 resources in many regions, but may also lead to a long, but initially hidden, legacy of
43 anthropogenic and climatic impacts on river flows and groundwater dependent ecosystems.

44

45

46 **Text**

47 Groundwater flow systems exist in dynamic balance with the climate, connecting interacting
48 zones of recharge (i.e. the replenishment of water in the subsurface) and discharge (the loss
49 of groundwater from the subsurface), with multiple feedbacks. As climate varies, changes in
50 the quantity and location of natural groundwater recharge lead to changes in groundwater
51 storage, water table elevations and groundwater discharge¹. These changes in time and space
52 play a central role in controlling the exchange of moisture and energy across the Earth's land
53 surface^{5,6} and connect processes critical to, for example, hydro-ecology, as well as carbon and
54 nutrient cycling⁷. Climate-groundwater interactions may also have played a key role in the
55 evolution of our own and other species⁸ and continue to be critical in setting the availability
56 of water for abstraction by humans in coupled food-water-energy systems¹. Recent global
57 mapping of water table depths⁹ and the critical zone¹⁰ suggest where interactions of climate
58 and groundwater may be most tightly coupled. However, they do not resolve where
59 groundwater systems are most sensitive to changes in climate and vice versa, or the
60 timescales over which such changes may occur.

61 Here, we derive and combine global scale analytical groundwater model results and other
62 hydrologic data sets to provide the first global assessment of the sensitivity of groundwater
63 systems to changes in recharge in both space and time (Figure 1), and discuss their utility as
64 an emergent constraint in understanding and modelling groundwater interactions with climate
65 and other Earth systems at the global scale.

66 We have characterized the mode of groundwater-climate interactions as being either
67 principally bi-directional or uni-directional using an improved formulation of the water table
68 ratio (*WTR*)^{11,12} mapped globally at high resolution (Figure 1a, Figure S1-2). The *WTR* is a
69 measure of the relative fullness of the subsurface and thus the extent of the water table's
70 interactions with topography. Values of *WTR*>1 indicate a topographic control on water table

71 conditions broadly correlating to shallow (<10 metres below ground level, m bgl) water table
72 depths (*WTDs*) globally (see Methods and Figure S3). This is indicative of a prevalently bi-
73 directional mode of groundwater-climate interaction (Figure 1c) where the climate system
74 can both give to the groundwater system in the form of recharge, and receive moisture back
75 via evapotranspiration if *WTDs* are shallow enough.

76 The land surface in such regions rejects a proportion of the potential recharge, and
77 groundwater can have a limiting control on land-atmosphere energy exchanges⁵; a tight two-
78 way coupling between groundwater and surface water is also common. In contrast, in
79 ‘recharge controlled’ areas where $WTR < 1$, water tables are more disconnected from the
80 topography and, while groundwater may still receive recharge from the land-surface, the
81 extent of two-way interaction between climate and groundwater is limited and the mode of
82 interaction is predominantly uni-directional (Fig. 1c).

83 We find that regions where $WTR > 1$ cover around 46% of the Earth’s land area (see Methods,
84 Figure 1a,b) and contribute to the large, but until recently underestimated, extent of
85 groundwater-vegetation interactions globally^{10,13,14}. Consistent with previous regional
86 analyses and the form of the governing equation (see Methods), our results indicate that bi-
87 directional interactions are more likely to occur in areas with high humidity, subdued
88 topography and/or low permeability. In contrast, regions with $WTR < 1$ are more common in
89 drier climates or more mountainous topography¹¹.

90 In order to assess the large scale temporal sensitivity of climate-groundwater interactions we
91 have used an analytical groundwater solution to quantify groundwater response times (*GRTs*)
92 globally and at high resolution. *GRT* is a measure of the time it takes a groundwater system to
93 re-equilibrate to a change in hydraulic boundary conditions¹⁵. For example, the *GRT*
94 estimates the time to reach an equilibrium in baseflow to streams (or other boundaries) after a
95 change in recharge rate, potentially from climate or land use change. Our results indicate that

groundwater often has a very long hydraulic memory with a global median *GRT* of nearly 6000 yrs, or approximately 1200 yrs when hyper-arid regions, where recharge is <5 mm/y, are excluded (Figure 1d,e). Only 25% of Earth's land surface area has response times of less than 100 yrs (herein called 'human timescale'). However, this is equivalent to nearly 44% of global groundwater recharge flux, calculated by aggregating contemporary recharge over the land area where *GRT*<100 y, expressed as a proportion of the total global recharge. Around 21% by area have uni-directional climate-groundwater interactions and response times on human timescales, mostly associated with high permeability geology suggesting a strong lithological control (Figure 2a).

The remainder (4%) in areas with bi-directional climate-groundwater interactions are mostly located in the humid, lowland, tropical regions with unconsolidated sediments (e.g. Amazon and Congo Basins, Indonesia), low-lying coastal areas (e.g. Florida Everglades, Asian megadeltas) or in high latitude, low topography humid settings (e.g. northeastern Canada, parts of northern Europe).

A powerful advantage of using analytical groundwater equations such as the *WTR* is that they allow us to directly assess the spatial sensitivity of the mode of climate-groundwater interactions. By taking the derivative of *WTR* with respect to recharge (Figure S4) we have a measure of the sensitivity of the relative fullness of the subsurface to changes in recharge (see Methods). Our results indicate that the mode of climate-groundwater interaction is very insensitive to relative changes in recharge (Figure 2b, Figure S5), with only 5% of the Earth's land surface switching mode for a 50% relative change in recharge rate. This represents a large change in natural groundwater recharge in the context of projections for the coming century¹⁶. However, when absolute recharge rate changes are considered, more sensitivity is apparent and a pattern emerges (Figure S6-7) that indicates the strong inverse relationship between the spatial and temporal sensitivity of groundwater systems to changes in recharge

that we observe (Figure 3b). At small, local scales our calculations may have relatively large uncertainties, stemming from the uncertainties in global data sets used for the analysis particularly for hydraulic conductivity (see Methods). However, at the larger scales considered here, Monte Carlo Experiments (MCE) indicate that, once the variance in each parameter is combined, the global estimates have relatively small standard deviations (Figures 1-2, S2).

The global pattern of *GRT* (Figure 1d) indicates a propensity for longer hydraulic memory in more arid areas. Despite the expected scatter due to geomorphological and lithological heterogeneity, there is a power law relationship between median *GRT* and groundwater recharge (*R*) such that $GRT \propto 1/R^y$ with $y \sim 2$ (Figure 3a). This discovery is not directly expected from the form of the governing equations but is rather an emergent property of groundwater system interactions with the Earth's land surface and climate system. The principal control on the observed power law is the distribution of perennial streams (Figure S8) to which the *GRT* is most sensitive, and which itself is strongly controlled by climate (Figure S9-11). How to characterize, quantitatively, this climatic control on the perennial stream distributions is a pertinent question for further hydro-geomorphological research.

We should not therefore expect *GRTs* to be static nor consider them as 'time constants' despite being mathematically equivalent to other diffusion processes. Rather, *GRTs* will evolve in time as both climate and geology vary the geometry and hydraulic properties of groundwater flow systems. This will occur over long but diverse timescales associated with changing river geometries.

Despite its importance, most global climate, Earth system, land surface and global hydrology models exclude groundwater or do not allow groundwater to flow between model grid cells¹⁸⁻²⁰. While our results suggest that the spatial distribution of the mode of climate-groundwater

interactions may be rather static over century long timescales, we have shown that nearly a half of the world's groundwater flux is responsive on 100 y timescales. Hence in order to capture the important mass and energy transfers correctly, which may affect regional precipitation and temperature dynamics^{5,6}, lateral flow circulation of groundwater must be incorporated into the next generation of global models rather than assuming within-grid-cell hydrological closure of the water budget as is currently often assumed²¹⁻²³. Our *GRT* calculations provide direct estimates of spin-up times to improve groundwater-enabled global models, without having to use the currently employed methods of extrapolation²². Given the long *GRTs* present over much of the Earth's land-surface, defining initial conditions with an equilibrium water table calculated for present-day climate conveniently, but wrongly, assumes stationarity in groundwater levels and fluxes. Since groundwater is known to be the part of the hydrological system that takes longest to achieve equilibrium²⁴, new approaches that incorporate the existence of long term transience should continue to be developed²⁵.

The global distribution of *GRTs* suggests that widespread, long-term transience in groundwater systems persists in the present day due to climate variability since at least the late Pleistocene in many semi-arid to arid regions (Figure 3a). This is consistent with observations of larger than expected groundwater gradients, given the current low recharge, that have been observed in present day arid zones²⁵. While groundwater residence time and groundwater response time are fundamentally different concepts, we also note the correspondence between high *GRT* and significant volumes of fossil-aged groundwater storage in arid regions^{2,26}. The outcome of this result is that groundwater discharge to oases, rivers or wetlands in otherwise dry landscapes will be particularly intransient in comparison to climate change, in as much as climate controls the variations in groundwater recharge. However, our results also indicate that groundwater response times tend to be greater in regions where water tables are most sensitive to changes in recharge (Figure 3b). This

follows from the fact that both the groundwater response time and the derivative of the water table ratio share a strong dependence on the square of the distance between perennial streams (L , compare Equations 10 and 14).

Away from these more arid contexts, the responsiveness of groundwater systems has recently been demonstrated to be as important as climate controls for the development of hydrological drought²⁷. For example, low *GRT* systems tend to enhance the speed of propagation of meteorological drought through to hydrological drought whereas higher *GRT* systems attenuate climate signals to a greater extent but also show greater lags in recovery from drought. Thus, even within relatively small geographic areas, geological variations can lead to very different drought responses even under similar climate variability. By way of a specific example, increasing lags between meteorological and hydrological drought indicators have been observed between the two most significant aquifers in the UK²⁸ in a manner consistent with what would be expected from our estimates of *GRT* (i.e. Cretaceous Chalk limestone - *GRTs* of months to years, Permo-Triassic sandstone – *GRTs* of years to 100s years, Figure 1d).

Our analysis therefore provides a new framework for understanding global water availability changes under climate change. First, the discovery of a power law relating groundwater recharge and *GRT* suggests that important areas of groundwater discharge in naturally water scarce parts of the world are likely to be more resilient to climate fluctuations than humid areas. However, where groundwater response times are higher, water tables also tend to be most sensitive to changes in recharge in the long term. Hence, accounting appropriately for groundwater-climate interactions within analyses of global water scarcity in the context of climate change is thus of great importance when explicitly considering the contribution of groundwater storage changes²⁹. Second, the long memory of groundwater systems in drylands also means that abrupt (in geological terms) changes in recharge or widely

distributed groundwater abstraction will leave longer legacies. There may also be initially ‘hidden’ impacts on the future of environmental flows required to sustain streams and wetlands in these regions. It is critical therefore that climate change adaptation strategies which shift reliance to groundwater¹ in preference to surface water should also take account of lags in groundwater hydrology³⁰ and include appropriately long timescale planning horizons for water resource decision making. Third, robust assessments of the impact of climate change on hydrological drought require estimates of ‘groundwater responsiveness’²⁷. The timescale of such responses can be directly informed by our results and improve the decision making process with regard to adaptation strategies to changing drought frequencies under climate change.

Figure Captions

Figure 1. Global distributions of water table ratios (*WTR*) and groundwater response times (*GRT*) with their conceptual interpretation as metrics of climate-groundwater interactions. (a) Global map of $\log(WTR)$ with hyper-arid regions of recharge (R) < 5 mm/y shaded grey¹⁷. **(b)** Frequency distribution of global values of $\log(WTR)$. **(c)** Conceptual model for *WTR* as a metric for either bi-directional or uni-directional groundwater-climate interactions - *WTR* is dependent on R , terrain rise (d), distance between perennial streams (L) and the saturated thickness of the aquifer (b). **(d)** Global map of $\log(GRT)$. **(e)** Frequency distribution of global values of *GRT* - median 5727 yrs (standard deviation, σ = 376 yrs), or 1238 yrs (σ = 92 yrs) when hyper-arid regions are excluded. **(f)** Conceptual model of *GRT* as a metric of the temporal sensitivity of groundwater-climate interactions.

Figure 2. Global distributions of the temporal and spatial sensitivity of the mode of climate-groundwater interactions. (a) Temporal sensitivity: percentage of uni-directional and bi-directional groundwater systems, by area globally, that will re-equilibrate significantly to changes in recharge on the timescale of <100 y or >100 y. **(b)** Spatial sensitivity: percentage of the global area that would change mode from bi-directional to uni-directional climate-groundwater interactions, or vice versa, for a relative change of 50% in recharge, given an unlimited amount of time. Mapped values use the baseline parameter set (see Methods). The median percentage

coverage of Earth's landmass for each category from the Monte Carlo Experiments is labelled in the key with standard deviations in percentage coverage shown in brackets. Grey areas represent contemporary recharge <5 mm/y (ref⁴⁷).

Figure 3. Global quantitative inter-relationships between climate and the temporal (*GRT*) and spatial (*WTR*) sensitivity of groundwater-climate interactions. (a) Globally, median *GRT* values scale approximately with the inverse of recharge (*R*) squared. Relationships between recharge and aridity index categories are shown on the top axes as derived in Figure S12. Box extents are at 25-75% percentiles, with Tukey whiskers and outliers. Histograms within each box represent median *GRT* values from each MCE realisation. **(b)** The sensitivity of climate-groundwater interactions in time (*GRT*) and space (*dWTR/dR*) are log-correlated. Each point uses median values for a geographic location from the MCE realisations. Inset plots are frequency distributions of the slope and r^2 derived from linear regressions carried out for each realisation indicating consistency in the relationship across the uncertainty range.

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Author Contributions

322 The idea for the paper was conceived by MOC and TG. Analyses were by all authors. The
323 manuscript was written by MOC with input from all authors.

324

325 **Data Availability statement**

326 Once the paper is accepted the main data outputs (i.e. *WTR* and *GRT* digital maps) will be
327 made freely available for download via an online data repository and a link included in the
328 final published version of the paper.

329

330

331 METHODS

332 Derivation of Equations

333 *Governing groundwater flow equations*

334 The governing equations were formulated by considering an ideal homogeneous, horizontal
335 unconfined aquifer bounded at one end ($x = L/2$) by a stream assumed to be a constant head
336 boundary and at the other ($x = 0$) by a no-flow boundary representing a flow divide
337 (Figure S13). The one-dimensional (Boussinesq) equation of groundwater flow for such an
338 aquifer receiving homogeneous recharge can be given as follows:

$$339 \quad \frac{\partial}{\partial x} \left(Kh \frac{\partial h}{\partial x} \right) = S \frac{\partial h}{\partial t} - R(t) \quad (1)$$

340 where K is hydraulic conductivity [LT^{-1}], S is storativity [-], $h(x,t)$ is hydraulic head [L], t is
341 time [T], x is distance [L] and $R(t)$ is groundwater recharge [LT^{-1}].

342 If changes in transmissivity due to fluctuations in groundwater heads are assumed to be
343 negligible, Equation (1) may be linearised as follows:

$$344 \quad T \frac{\partial^2 h}{\partial x^2} = S \frac{\partial h}{\partial t} - R(t) \quad (2)$$

345 where T is transmissivity [L^2T^{-1}], and $T = KH$, with H the average saturated thickness [L].

346 The lateral boundary conditions are as follows:

$$347 \quad \frac{\partial h(0,t)}{\partial x} = 0, \quad h\left(\frac{L}{2}, t\right) = b \quad (3)$$

348 The parameter L is thus a characteristic length equivalent to the distance between perennial
349 streams which act as fixed head groundwater discharge boundaries.

350 *Water table ratio (WTR) derivations*

351 For steady state flow, where $h(x,t)$ becomes $h(x)$, the solution to Equation 1 for the stated
 352 boundary conditions is:

$$353 \quad h(x) = \left(b^2 + \frac{R}{K} \left(\frac{L^2}{4} - x^2 \right) \right)^{0.5} \quad (4)$$

354 At the flow divide, $x = 0$, therefore:

$$355 \quad h(0) = \sqrt{b^2 + \frac{RL^2}{4K}} \quad (5)$$

356 For steady state flow, the solution to the linearised form, Equation 2, for the stated boundary
 357 conditions is:

$$358 \quad h(x) = \frac{R}{2T} \left(\frac{L^2}{4} - x^2 \right) + b \quad (6)$$

359 At the flow divide, $x = 0$, therefore:

$$360 \quad h(0) = \frac{RL^2}{8T} + b \quad (7)$$

361 The *WTR* is defined¹² as the ratio of the head at the flow divide above the fixed head
 362 boundary (i.e. $h_0 - b$) to the maximum terrain rise above the fixed head boundary, d [L].

363 This yields a new, non-linearised, form of the *WTR*, from Equation 5 as follows:

$$364 \quad WTR_{NL} = \frac{\sqrt{b^2 + \frac{RL^2}{4K}} - b}{d} \quad (8)$$

365 For the linearised form, from Equation 7, and as originally given by ref¹², the *WTR* is:

$$366 \quad WTR_L = \frac{RL^2}{8Td} = \frac{RL^2}{8KHd} \quad (9)$$

367 Equations 8 and 9 become equivalent for combinations of small L or R , or large K .

368 All maps and analysis presented in this paper use the non-linear form of the *WTR*
 369 (Equation 8) with the exception of Figure S1 where the two versions are compared, and

calculated using the L parameters derived using a minimum river discharge threshold of $0.1 \text{ m}^3/\text{s}$. A comparison of global maps and frequency distributions for the linear and non-linear forms are shown in Figure S1-2. The frequency distribution comparison (Figure S2) shows that the new non-linear formulation has a narrower and more symmetric distribution with a median closer to zero than the linearised form. This is indicative of its better physical representation such that the extent of higher WTR s is limited by the feedback between higher water table elevation and concomitant increases in transmissivity inherent in the non-linear Boussinesq equation (Equation 1).

The WTR is a measure of the relative fullness of the subsurface and thus the extent of the water table's interactions with topography. We have therefore used the WTR to characterize the dominant mode of groundwater-climate interactions as being either principally bi-directional or uni-directional based on whether they are 'topographically controlled' ($WTR > 1$) or 'recharge controlled' ($WTR < 1$), respectively. This is a reasonable approximation since a global comparison with water table depths (WTD s) (Figure S3) indicates that $WTR > 1$ broadly correlates to shallow (<10 metres below ground level) water table conditions. This condition is indicative of a prevalently bi-directional mode of groundwater-climate interaction where the climate system can both give to the groundwater system in the form of recharge, and receive moisture back where local variations in WTD s enable evapotranspiration to occur from groundwater directly. In contrast, areas with $WTR < 1$ show increasingly large WTD s well beyond plant rooting depths leading to predominantly uni-directional climate-groundwater interactions where the groundwater system receives recharge from the climate system but there is more limited potential for feedback in the other direction. The sensitivity of the WTR to changing recharge is given by differentiating Equation 8 with respect to R :

$$\frac{dWTR_{NL}}{dR} = \frac{L^2}{8Kd} \left(b^2 + \frac{RL^2}{4K} \right)^{-0.5} \quad (10)$$

This equation represents the sensitivity of the maximum head to recharge relative to the topography which can be understood as the sensitivity of the ‘fullness’ of the subsurface to changes in recharge.

Following from Equations 8, we calculate the recharge required for the WTR to equal 1 for every grid cell as:

$$R_{WTR=1} = \frac{4K}{L^2} (d^2 + 2db) \quad (11)$$

The difference between R and the values given in Equation 11 then gives an expression for the change in recharge (ΔR) needed to effect a change in the WTR across the transition between topography control (bi-directional climate-groundwater interactions) and recharge control (unidirectional climate-groundwater interactions) modes. In absolute terms this is:

$$\Delta R_{abs} = R - R_{WTR=1} \quad (12)$$

and in relative terms it becomes:

$$\Delta R_{rel} = \frac{R - R_{WTR=1}}{R} \quad (13)$$

Groundwater response time (GRT) definition

The groundwater response time is, in general terms, a measure of the time it takes a groundwater system to respond significantly (as defined below) to a change in boundary conditions^{15,31-35} and is defined here as follows:

$$GRT = \frac{L^2 S}{\beta T} \quad (14)$$

where β is a dimensionless constant, T is transmissivity [$L^2 T^{-1}$], S is storativity [-] and L is the distance between perennial streams [L]. To illustrate why this equation defines a time of

response consider a groundwater mound such as that shown in Figure S13. Let the initial shape of the mound (of maximum height A), due to some steady recharge, be given by:

$$h(x, 0) = A \cdot \cos\left(\frac{\pi x}{L}\right) \quad (15)$$

If recharge suddenly ceases (i.e., a step change) then it can be shown, in the manner of ref³³, that the solution to the linearised Equation 2 without recharge (i.e. $R(t)=0$) is:

$$h(x, t) = h(x, 0) \cdot \exp\left(-\frac{t}{GRT}\right) \quad (16)$$

for β is equal to π^2 .

Hence, for this case, the GRT controls the timescale for the groundwater levels to decay exponentially to reach 63% re-equilibrium after a change in boundary (recharge) conditions (i.e., an “e-folding” timescale). This value for β was chosen in order to be consistent with mathematically equivalent uses of ‘time constants’ (often denoted as τ), in other branches of science.

As outlined by ref³⁴, comparing the timescale of a particular forcing to the GRT can be a useful measure of the degree of transience a groundwater system will manifest in terms of variations in lateral groundwater flow. However, there is an important difference to note in the case of a step change in conditions, as used to define GRT in Equation 14, in comparison with a periodic variation in the forcing recharge (of period P). For the step change case outlined above, both heads and fluxes decay exponentially after the change in recharge. However, in the periodic case, where $GRT \gg P$, variations in recharge lead to very stable groundwater fluxes (including at the downstream lateral boundary) but large temporal changes in groundwater head across much of the aquifer³⁵. Thus, it is important to distinguish between the control of GRT on the degree of transience in either heads or fluxes, depending on the nature of the boundary conditions.

Spatial input data and manipulation

Global mapping of the distance between perennial streams (L)

The distance between perennial streams (L) was calculated using a globally consistent river network provided by the HydroSHEDS database³⁶ which was derived from the 90 m digital elevation model of the Shuttle Radar Topography Mission (SRTM). For this study, we extracted the global river network from the HydroSHEDS drainage direction grid at 500 m pixel resolution by defining streams as all pixels that exceed a long-term average natural discharge threshold of 0.1 cubic meters per second, resulting in a total global river length of 29.4 million kilometers. Smaller rivers with flows below this threshold were excluded as they are impaired by increasing uncertainties in the underpinning data. However, the sensitivities of the most important results of this paper to the chosen threshold are considered in our uncertainty analysis below. Estimates of long-term (1971-2000) discharge averages have been derived through a geospatial downscaling procedure³⁷ from the 0.5° resolution runoff and discharge layers of the global WaterGAP model (version 2.2, 2014) a well-documented and validated integrated water balance model^{16,38}. Only perennial rivers were included in the assessment; intermittent and ephemeral rivers were identified through statistical discharge analysis (lowest month of long-term climatology is 0) and extensive manual corrections against paper maps, atlases and auxiliary data, including the digital map repository of National Geographic³⁹. L was calculated for every pixel of the landscape (Figure S8) by identifying the shortest combined Euclidean (straight-line) distance between two river locations at opposing sides of the pixel. Neighbourhood low pass filters (5x5 kernel size) were applied to remove outlier pixels and speckling. All calculations were performed in ESRI© ArcGIS environment using custom-made scripts.

Global mapping of the water table ratio (WTR), groundwater response times (GRT) and other expressions

463 Global *WTR* maps were created from the above equations using: the recharge rate (R in m/y),
464 based on ref¹⁷, a minimum saturated thickness of the aquifer (b) set to 100 m (refs^{40,41}), the
465 distance between two perennial streams (L , in m, as described above), intrinsic permeability
466 values (m^2) reported in ref⁴⁰ were converted to hydraulic conductivity (m/s) by assuming
467 standard temperature and pressure (1×10^7 multiplication factor) and then converted to units
468 of m/y. The maximum terrain rise between rivers (d , in m) was based on the range of
469 elevations in the 250m GMTED2010 data set⁴².

470 The *GRT* was mapped using the same L data and hydraulic conductivity values as for the
471 *WTR* calculations. Transmissivity (T , m^2/y) was calculated by multiplying the hydraulic
472 conductivity with a fixed saturated thickness of 100 m (refs^{40,41}). It was assumed that
473 storativity (S) for unconfined aquifers is dominated by the specific yield and that this can be
474 approximated by mapped porosity values⁴⁵. Owing to the significant uncertainties in these
475 assumptions for calculating T and S values the parameters were subjected to a Monte Carlo
476 analysis as described below.

477 Each of the data sets was prepared to match a global equal-area projection with a grid size of
478 1 km x 1 km, and the calculations of the data sets were performed in ArcGIS. To avoid
479 mathematical problems, for zero values of d and R , 1 and 0.00001 were added, respectively.
480 For *WTR* estimates, regions where contemporary groundwater recharge was estimated as < 5
481 mm/y (ref¹⁷) were excluded from the analysis due to the increasingly large relative
482 uncertainties in recharge below this range, and the resulting unrealistic sensitivity of the
483 resulting *WTR* estimates. For deriving the frequency distributions and comparisons of
484 parameters from the range of derived geo-spatial data sets, point values were taken from each
485 raster of interest for 10,000 randomly distributed locations across the Earth's land-surface.
486 Global distributions of the parameters d , K and S are given in Figure S10 and relationships

between R and L ; d and WTR ; and R and WTR are explored in Figure S9, and Figures S11, respectively. All areal calculations ignore the Antarctic landmass.

Although we have made best use of coherent available global datasets at high (1 km) resolution for the calculations, our results are intended for appropriate large scale interpretation, not detailed local analysis.

Justification of the model assumptions

Our calculations are based on mapped surface lithology only and, as such, they represent a first estimate of the response of unconfined groundwater across the global land surface. The more complex responses of regional or local confined aquifers, which may be locally important to discerning groundwater-climate interactions, are not considered. However, such confined aquifers only cover around 6-20% of the Earth's surface⁴³, are often located in more arid parts of the world and are, by definition, inherently less connected to the land surface and climate-related processes.

Using 1-D analytical solutions to the groundwater flow equations gives a powerful advantage over the use of more complex models in enabling the sensitivity of the key parameters controlling patterns and timescales of climate-groundwater interactions to be analysed analytically. This, for example, allows us to sample the entire parameter space directly rather than a restricted subset via a limited ensemble of more computationally expensive numerical model runs. Equation 1 assumes the validity of the Dupuit-Forchheimer approximation whereby the water table is assumed to be a true free surface governed by effective hydraulic parameters and that water pressure in the direction normal to the flow is approximately hydrostatic. This is a good approximation when the ratio of the lateral extent of the average saturated depth is more than approximately 5 times its depth¹², i.e. $H/(L/2) < 0.2$ (see Figure S13). Calculating the maximum saturated depth h_{max} as the smaller of $d+b$ or h_0 , and

511 approximating the average saturated depth as $(h_{max} + b)/2$, we find that the criterion
512 $H/(L/2) < 0.2$ is met in 96% of our global grid calculations. Locations which fail this test are
513 all in mountainous regions where Equation 1 cannot account accurately for steep hillslope
514 groundwater hydraulics and hence our results may be less reliable in such areas.

515 The *GRT* is a parameter which consistently appears in solutions to the groundwater flow
516 equations and has been used for decades³² as a robust estimate for the timescale of re-
517 equilibration of a groundwater system following a change in boundary conditions^{8,15,30-35,44-48}.
518 Thus it is an appropriate metric for long term transience which is currently impossible to
519 model in state of the art coupled groundwater-surface water models, which are limited to
520 short run times even for regional scale analyses due to their massive computational demands.
521 More realistic aquifer geometries and initial water table configurations lead to behaviours
522 which are more complex than the case of a simple exponential decay⁴⁶, and non-uniform flow
523 fields (strong convergence or divergence) can also lead to variations in *GRT* (refs^{44,47,48}). We
524 have therefore included these factors in an uncertainty analysis as outlined below.

525 While the models used here cannot represent the detailed process interactions in the way that
526 a distributed fully coupled 3-D model would, they have a strong theoretical basis and show
527 consistency with other large scale studies based on very different model assumptions and data
528 sets. Justification for the approach of using *WTR* as a proxy for the mode of climate-
529 groundwater interaction is given in at least four ways. First, at global scale, similarities of
530 *WTR* with shallow *WTD* globally⁹ are strong (Figure S3), given the very different model
531 assumptions and data sets employed in the two studies. Second, at a continental scale for the
532 contiguous US a recent study compared the results of a physically based, 3-D, fully coupled
533 surface water-groundwater model validated against water table depth data, against the *WTR*
534 metric⁴¹. The results show scatter as expected due to variations in the derivation of the
535 comparative characteristic length scales used in the comparison. However, general trends and

geographic patterns at a regional scale compare well between the WTD computed by the fully coupled model and the calculated WTRs. Third, also at a continental scale for the contiguous US, a systematic relationship has been shown between *WTR* and mean stream junction angles which are indicative of a strong coupling between surface and subsurface⁴⁹. Lastly, comparisons of *WTR* calculations against a more complex 3-D regional groundwater flow model, has indicated that the *WTR* is a robust indicator of groundwater's connection to the land surface as it is a strong predictor of the propensity for local versus regional flow conditions⁵⁰. Our analyses thus allow us to make a robust first global scale estimate of the sensitivity of climate-groundwater interactions, while enabling the range of uncertainty to be fully and directly appreciated.

Uncertainties and Monte Carlo experiments

We ran 10,000 Monte Carlo experiments (MCE) at 10,000 randomly distributed locations across the Earth's land-surface to investigate the range of uncertainty due to parameter uncertainties as well as model structural simplifications.

Hydraulic conductivity (*K*) was allowed to vary log-normally within uncertainty ranges defined in refs^{40,50}, this parameter having by far the highest uncertainty of any others used in our calculations. Groundwater recharge (*R*) values were taken from ref¹⁷ but allowed to vary through a normal distribution with a standard deviation of 22% of this baseline, chosen according to the difference with a contrasting global recharge distribution^{52,53} commonly used in other global hydrological calculations. Storativity (*S*) was sampled from a normal distribution with standard deviations of 25% of the mapped value⁵³. Although the absolute error in the DEM used is only 1-2 m, we allowed the maximum terrain rise (*d*) to vary normally with a standard deviation of 10% to allow for uncertainties due to gridding. The minimum saturated thickness of the aquifer (*b*) was allowed to vary log-normally around

100 m with a standard deviation of 0.3 orders of magnitude. Sampled distributions were cut off at zero to stop meaningless negatives being included in the calculations.

Parameter uncertainty in the distance between perennial streams (L), calculated from the variation in L for an order of magnitude change in discharge threshold used to define the stream network (from 0.1 to 1 m³/s), gives a median uncertainty of a factor of 1.9. However, there is also additional uncertainty to L due to the choice of the one-dimensional groundwater flow solutions applied, which ignore non-uniform (i.e. convergent or divergent) flow fields which are common in real catchments. In order to account for the maximum likely range of possible uncertainty, we have compared the 1-D analytical solutions used here to cases of radial flow which represent an extreme 2-D non-uniform flow end-member for natural groundwater flow systems. By equating the distance between perennial streams (L) to be equal to the radius of the flow domain for the equivalent radial solutions, we can estimate the impact of this choice on both WTR and GRT . For WTR , by replacing Equation 6 with Eq 30.11 from ref⁵⁴, the average error is approximately a factor of 2. For the GRT , comparison of recession timescales for 1-D and radial flow cases (e.g. Appendix A of ref⁴⁶) indicates a similar level of uncertainty due to non-uniform flow as for the WTR . We therefore added a log-normal variation in L with a standard deviation of 0.3 orders of magnitude to accommodate the likely range of combined parameter and structural uncertainty.

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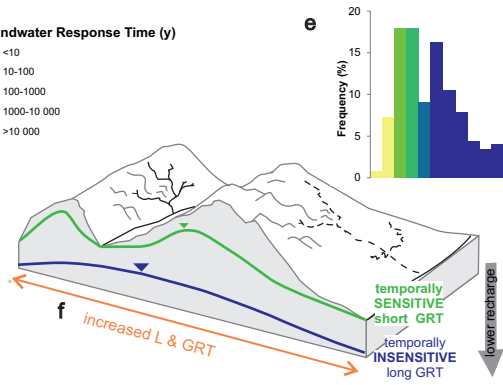
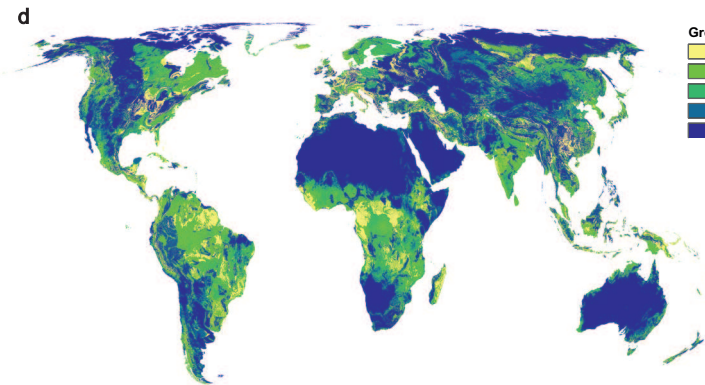
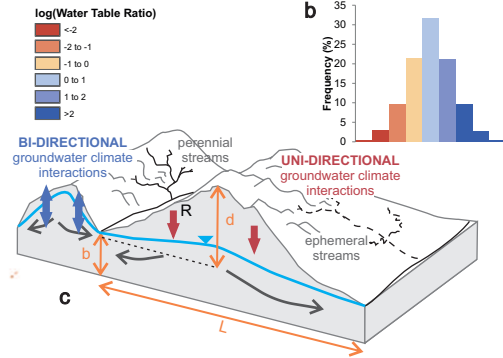
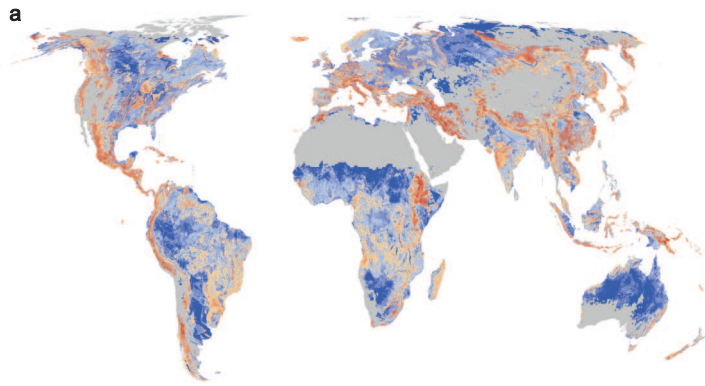
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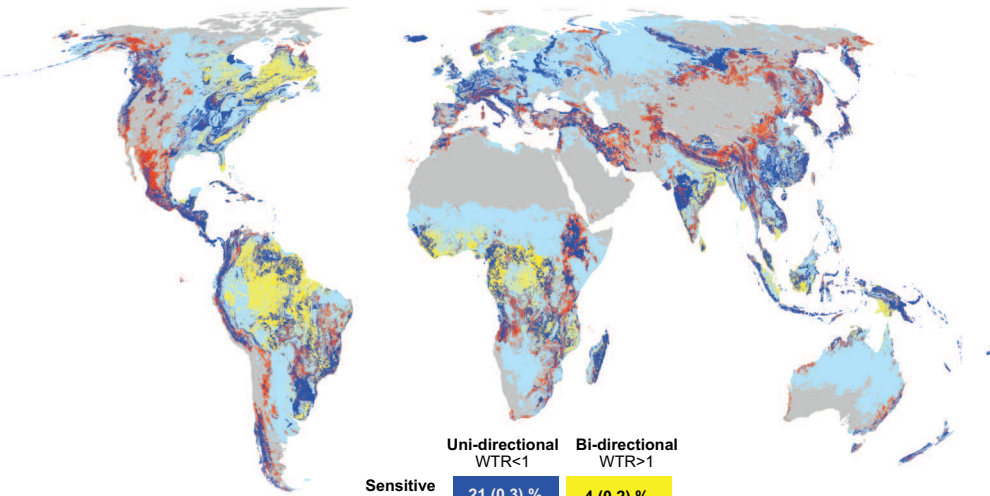
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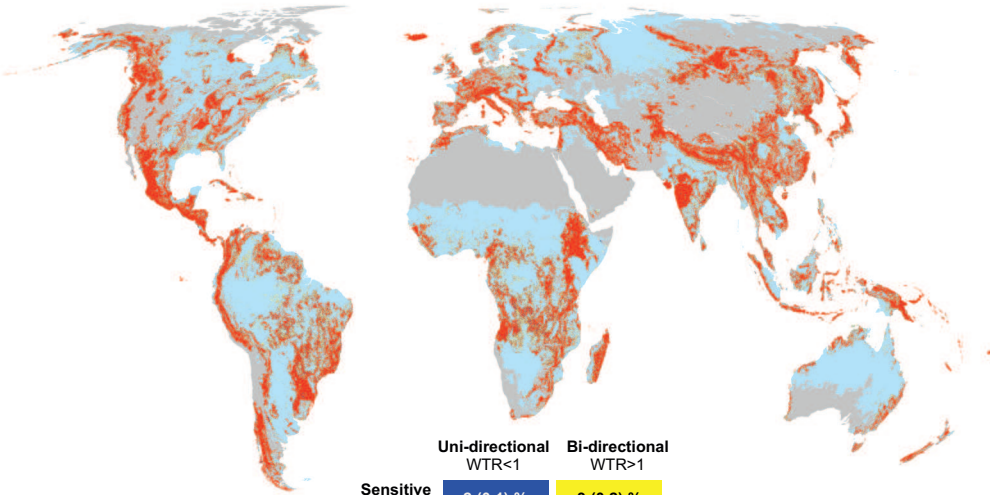


a



	Uni-directional WTR<1	Bi-directional WTR>1	
Sensitive GRT<100 y	21 (0.3) %	4 (0.2) %	R<5 mm/y 26 (0.1) %
Insensitive GRT>100 y	7 (0.2) %	42 (0.4) %	

b



	Uni-directional WTR<1	Bi-directional WTR>1	
Sensitive $\Delta R < 50\%$	2 (0.1) %	3 (0.2) %	R<5 mm/y 26 (0.1) %
Insensitive $\Delta R > 50\%$	26 (0.3) %	42 (0.3) %	

